

Chemical composition of Earth, Venus, and Mercury

(planets/solar nebula/element abundances/mantle/core)

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ABSTRACT Model compositions of Earth, Venus, and Mercury are calculated from the premise that planets and chondrites underwent four identical fractionation processes in the solar nebula. Because elements of similar properties stay together in these processes, five constraints suffice to define the composition of a planet: mass of the core, abundance of U, and the ratios K/U, Ti/U, and FeO/(FeO + MgO). Complete abundance tables, and normative mineralogies, are given for all three planets. Review of available data shows only a few gross trends for the inner planets: FeO decreases with heliocentric distance, whereas volatiles are depleted and refractories are enriched in the smaller planets.

It has been known for over a century that the inner planets differ in density and, hence, in composition (1, 2). But a single characteristic, such as density, cannot constrain the abundance of 83 stable elements without additional assumptions. One fruitful assumption (3, 4) is that the planets formed by exactly the same processes as the chondrites, both being condensates from the solar nebula that experienced the same few fractionation processes (5). In chondrites, elements of similar cosmochemical properties are fractionated by similar factors, and they divide into five groups on this basis. If this is also true for planets, then we only require the abundances of five "index elements"—one for each group—in the planet to calculate the abundances of all 83 elements.

Compositions based on this approach have been calculated for Earth (4), Moon (4, 6), Mars (7, 8), and the eucrite parent body—presumably the asteroid 4 Vesta (refs. 9, 10; see also refs. 11 and 12). All except those for the Earth have gone through two iterations. The present paper corrects this oversight and also gives tentative model compositions for Venus and Mercury, although index data for these two planets are still incomplete. Values for noble gases are tabulated but not discussed; they will be taken up in a future paper.

We must stress that the compositions calculated by this model are working hypotheses to be tested and refined, not definitive truths to be enshrined in handbooks. For the Earth and Moon, for which data are ample enough to permit numerous tests, it appears that the first-generation models (4) are reasonable first approximations although still subject to further improvement (13-20). The rules for such improvements are quite restrictive; if one element is changed, all other elements from that cosmochemical group must be changed by the same factor.

THE MODEL

The basic assumption is that planetary matter condensed from a cooling solar gas (2, 21, 22). Under conditions of thermodynamic equilibrium, three types of material condense: (i) an early condensate rich in Ca, Al, and other refractories, (ii) nickel-iron, and (iii) magnesium silicates (Fig. 1). Below ~700 K, metal and silicates take up volatile elements. In addition,

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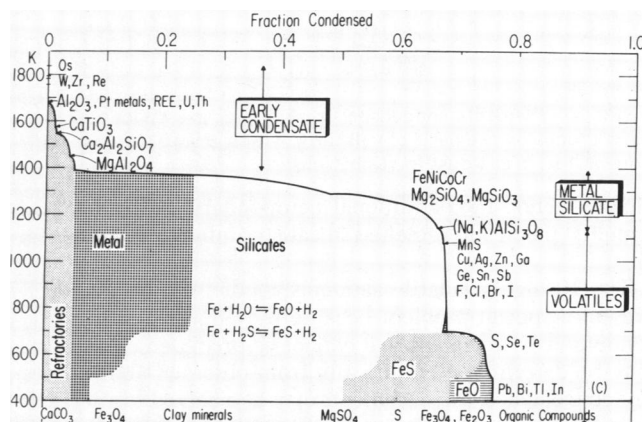


FIG. 1. Condensation of solar gas at 10^{-4} atm (10.1 Pa). Three types of dust condense from a solar gas: an early condensate consisting of refractory minerals, metallic nickel-iron, and magnesium silicates (21, 22). On cooling, iron reacts with H_2O and H_2S to give FeO and FeS. At this point, the mineral assemblage resembles that of the ordinary chondrites and inner planets. Major changes occur below 400 K, yielding a carbonaceous chondrite mineral assemblage.

metal reacts with H_2S and H_2O to form FeS (a fourth component) and FeO, which somehow entered the Mg silicates.

Judging from the chondrites, these primary condensates did not stay together in their original, cosmic proportions but became mutually fractionated by physical processes in the nebula (arrows in Fig. 2). The nature of these processes is not known, but likely possibilities are preferential settling of larger or earlier-condensed grains and preferential sticking of metal grains by ferromagnetic attraction (23, 24). Some part of the material also suffered brief remelting (chondrule formation) shortly before accretion, causing volatile loss and reduction of

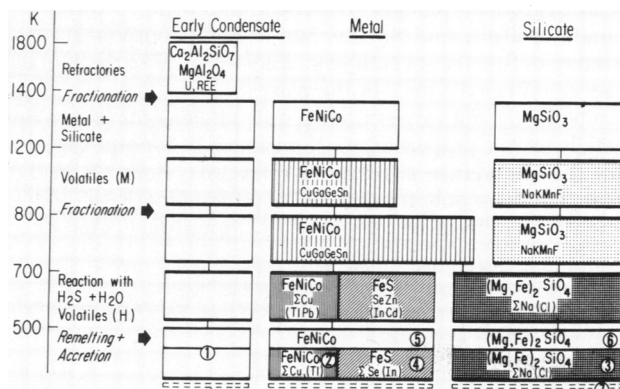


FIG. 2. Evolution of planetary matter in solar nebula. Three fractionation processes (arrows), not predicted by the condensation sequence, take place on cooling. They effectively double the number of components, from three to six. Presence of moderately (M) and highly (H) volatile elements is indicated by light and dark shading. Partially condensed elements are enclosed in parentheses; for convenience, they are assigned to a seventh component that is volatile-rich and resembles carbonaceous chondrites.

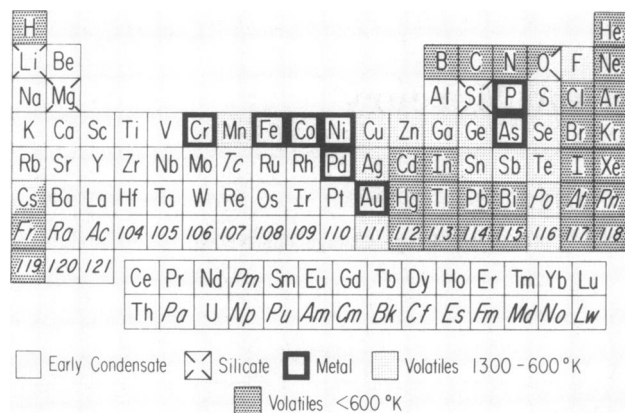


FIG. 3. Elements condensing from a solar gas divide into five major groups on the basis of volatility. These groups stay together, more or less, in gas/dust fractionations.

FeS to metal. Two additional components are thus made: remelted metal and remelted silicate (numbers 5 and 6 in Fig. 2).

The 19 highly volatile elements condensing below 600 K present a special problem because they may be only partly condensed, to an extent that cannot be predicted *a priori* but may be determined empirically from observed abundances. For convenience, these elements may be assigned to a seventh, volatile-rich component (no. 7 in Fig. 2) resembling some type of carbonaceous chondrite (4). [Actually, this "group" consists of three subgroups—volatile metals, major volatiles such as H, C, and N, and noble gases—that differ in condensation behavior and should really be represented by three different index elements (7, 8).]

Sulfur also is hard to constrain. Its cosmic abundance is comparable to that of Fe [5.0×10^5 vs. 8.9×10^5 atoms/ 10^6 Si atoms (4, 25)] and so it may not fully condense if less than 5×10^5 atoms of Fe^0 remain after metal loss and FeO formation. Hence, we normally base S on the cosmic S/K ratio or on the amount of available Fe, whichever is smaller (4, 6). However, these values may be too low; part of the S volatilized during chondrule formation may recondense because the back-reaction between Fe and H_2S is rather fast (21). At least for Earth, there is an additional clue: the density of the core implies the presence of $\approx 10\%$ of some light element such as S or O or both (26, 27). We have chosen a value of 9% S for Earth's core.

Although the model has seven components, it actually has only six degrees of freedom. The fraction remelted is the same for metal, sulfide, and silicate; and only six components are needed to define the seventh (e.g., silicate) because the sum of the seven equals the mass of the planet.

Each component carries its own suite of elements (Fig. 3) in generally cosmic proportions, so if we know U, for example, we know the abundance of 37 other refractory elements in the planet. Thus, we only need global abundances of five "index elements" to determine the planet's composition (Table 1).

Table 1. Index elements and ratios

Component	$T_{\text{cond}}, \text{K}$	Index	Earth	Venus*	Mercury*
Early condensate	>1400	U, ppb	14.3	15	11
Metal	1400–1300	Core, %	32.4	32	68
Unremelted	1300–600	K/U†	9440	10,000	2000
Volatile-rich	<600	Tl/U†	0.27	0.27	0.004
FeO	≈ 900 –500	f‡	0.12	0.07	0.03

* Poorly determined values are given in *italics*.

† Mass ratios.

‡ f = molar ratio, $\text{FeO}/(\text{FeO} + \text{MgO})$.

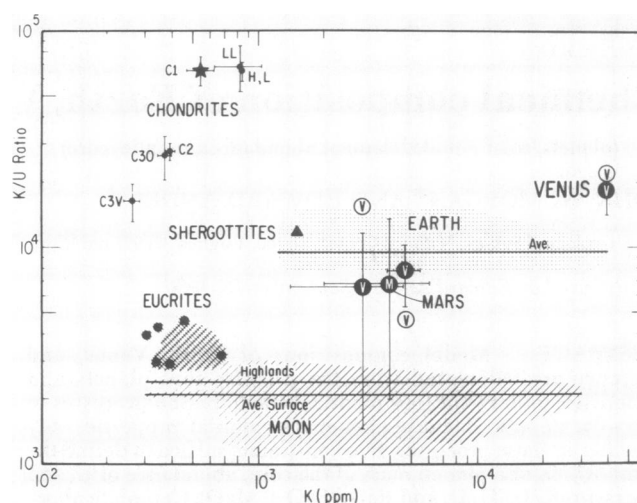


FIG. 4. K/U ratios of meteorites and planets. Several chondrite classes (30) have K/U ratios below the C1 chondrite value (star), reflecting depletion of K during chondrule formation (remelting). Ratios in planetary surface rocks (4, 31–33, †) and differentiated meteorites (30) are even lower. Because K and U do not fractionate readily in igneous processes, these ratios may be representative of the bulk planets. ●, K/U; ○, K/Th $\times 3.6$.

The first two, in principle, can be estimated directly from geophysical data (heat flow, density), with some iteration via thermal and seismic models. Fe can thus be determined to better than $\pm 10\%$ in all three planets, but U is known to only about $\pm 30\%$ in the Earth (28, 29) and not at all for Venus and Mercury; the (Earth-like and Moon-like) values in Table 1 are educated guesses, based on the apparent trend of U content with size (8) and a few scraps of evidence suggesting Earth-like and Moon-like composition for these two planets.

The next two index elements, K and Tl, can be estimated from their abundance ratios relative to U in surface rocks. These ratios tend to be nearly constant in a given planet (Fig. 4), reflecting the coherence of these large ions in igneous processes (3, 31). Three U values for Venus are known from γ -ray spectrometry by the spacecraft Venera 8, 9, and 10 (32, ‡); three additional values (open circles in Fig. 4) can be derived indirectly from the Th values by using the rather constant Th/U ratio in meteorites and many igneous rocks, 3.6. The single value for Mars is rather uncertain.† No data whatsoever are available for Mercury, and we have therefore used the average γ -ray spectrometric value for the surface of the Moon (33).

In the same spirit, we have chosen Earth-like and Moon-like Tl/U ratios for Venus and Mercury (Table 1), based mainly on the similar C and N contents of Earth and Venus (7, 34) and the distinctly lower volatile abundances of smaller planets such as Mars and the Moon (8). These values may be adequate for volatile trace metals and H, C, and N but not for noble gases.

The last index, FeO, measures the degree to which Fe became oxidized in the solar nebula (Figs. 1 and 2). In principle, it is given by the average $\text{FeO}/(\text{FeO} + \text{MgO})$ in the mantle silicates, but even for the Earth, this value is not directly determined because the lower mantle, of possibly higher FeO content (20), has not yet been sampled. Nonetheless, geophysical models point to a ratio of 0.12 ± 0.02 (35). Beyond the Earth, this ratio seems to increase with distance from the Sun, judging both from the abundance of Fe^{2+} in surface rocks (which is related to FeO content of the mantle) and from model compositions for Mars and the eucrite parent body (Fig. 5). The three classes of ordinary chondrites fit smoothly into this trend,

‡ Surkov, Yu. A. (1980) Paper presented at 23rd COSPAR Meeting, Budapest, Hungary, 2–14 June 1980.

Table 2. Elemental abundances in Earth, Venus, and Mercury

Element	Earth	Venus	Mercury	Element	Earth	Venus	Mercury
H, ppm	33	35	0.4	Ru, ppm	1.18	1.23	0.91
⁴ He*	111	11,000	—	Rh, ppb	252	265	194
Li, ppm	1.85	1.94	0.87	Pd, ppm	0.89	0.87	1.79
Be, ppb	45	47	34	Ag, ppb	44	49	7.2
B, ppb	9.6	10.0	0.11	Cd, ppb	16.4	17.2	0.19
C, ppm	446	468	5.1	In, ppb	2.14	2.24	0.024
N, ppm	4.1	4.3	0.046	Sn, ppb	390	430	64
O, %	30.12	30.90	14.44	Sb, ppb	35	39	5.7
F, ppm	13.5	15	2.2	Te, ppb	1490	830	122
²⁰ Ne*	0.50	49	—	I, ppb	13.6	14.3	0.16
Na, ppm	1250	1390	200	¹³² Xe*	0.0168	1.61	—
Mg, %	13.90	14.54	6.50	Cs, ppb	15.3	17.0	2.5
Al, %	1.41	1.48	1.08	Ba, ppm	4.0	4.2	3.1
Si, %	15.12	15.82	7.05	La, ppb	379	397	291
P, ppm	1920	1860	390	Ce, ppb	1010	1060	780
S, %	2.92	1.62	0.24	Pr, ppb	129	135	99
Cl, ppm	19.9	20.9	0.23	Nd, ppb	690	723	530
³⁶ Ar*	2.20	210	—	Sm, ppb	208	218	160
K, ppm	135	150	22	Eu, ppb	79	83	61
Ca, %	1.54	1.61	1.18	Gd, ppb	286	300	220
Sc, ppm	9.6	10.1	7.4	Tb, ppb	54	56	41
Ti, ppm	820	850	630	Dy, ppb	364	382	280
V, ppm	82	86	63	Ho, ppb	80	84	61
Cr, ppm	4120	4060	7180	Er, ppb	231	242	177
Mn, ppm	750	460	150	Tm, ppb	35	37	27
Fe, %	32.07	31.17	64.47	Yb, ppb	229	240	176
Co, ppm	840	820	1690	Lu, ppb	386	405	297
Ni, %	1.82	1.77	3.66	Hf, ppb	230	241	177
Cu, ppm	31	35	5.1	Ta, ppb	23.3	24.4	17.9
Zn, ppm	74	82	12.1	W, ppb	180	189	139
Ga, ppm	3.1	3.4	0.50	Re, ppb	60	64	46
Ge, ppm	7.6	8.4	1.24	Os, ppb	880	920	670
As, ppm	3.2	3.1	6.4	Ir, ppb	840	890	650
Se, ppm	9.6	5.4	0.79	Pt, ppm	1.67	1.76	1.29
Br, ppb	106	111	1.2	Au, ppb	257	250	516
⁸⁴ Kr*	0.0236	2.30	—	Hg, ppb	7.9	8.3	0.09
Rb, ppb	458	509	75	Tl, ppb	3.86	4.05	0.044
Sr, ppm	14.5	15.2	1.11	²⁰⁴ Pb, ppb	1.58	1.66	0.018
Y, ppm	2.62	2.74	2.01	Bi, ppb	2.94	3.08	0.034
Zr, ppm	7.2	7.5	5.5	Th, ppb	51.2	53.7	39.4
Nb, ppb	800	840	610	U, ppb	14.3	15.0	11.0
Mo, ppm	2.35	2.47	1.81				

Most values are based on the solar-system abundances of Cameron (25), with some later revisions (4). H, C, and N are based on their ratios to Ti in C3V chondrites (7). B and F (9 and 780 atoms per 10⁶ Si) data are from A.G.W. Cameron (personal communication), and Zr was changed to 12.9 atoms/10⁶ Si. Cs, an element lying between the two main groups of volatiles, was adjusted to the empirical Rb/Cs ratio in the Earth's crust, 30. Values in *italics* are poorly determined.

* Noble gas concentrations are given in units of 10⁻⁸ cm³/g. The values for Venus are scaled to the Pioneer Venus ³⁶Ar value and may well be too high by substantial factors.

if placed at arbitrary distances of 1.2–2 AU, the place of origin favored by some authors (36).

On the sunward side of the Earth, two data points are available. A spectrophotometric measurement for Mercury (37) suggests that pyroxenes on its surface have about the same FeO content as do those in lunar highland soils—i.e., 5.5%. This FeO must come from local rock, not from a meteoritic component, because the latter typically comprises less than 4% of a regolith and contains <30% FeO (38). For Venus, the latest thermodynamic calculations, assuming equilibrium between surface rocks and the atmosphere (39), suggest an FeO activity of 10⁻³–10⁻²—i.e., 1–2 orders of magnitude below the value for the Earth. But this result applies only to the surface rocks, whose chemistry is controlled by the massive, hot atmosphere. It does not preclude a higher FeO content in the bulk planet.

On the strength of the Mercury point—unaffected by an atmosphere—it appears that the FeO content of planetary matter falls much less steeply with heliocentric distance than

expected from the Fe/FeO equilibrium in Fig. 1 and an $a^{-0.5}$ dependence of T (dashed line in Fig. 5), let alone the still steeper T-dependence ($\approx a^{-1.1}$) favored by some authors (36). We have therefore chosen rather high FeO contents for Venus and Mercury (Table 1), based on the (tenuous!) assumption that the trend in Fig. 5 is monotonic.

It is an interesting question why the FeO content of planets varies so little with heliocentric distance (Fig. 5). One possibility is a very small temperature gradient in the solar nebula, smaller than any considered in theoretical models. Such a flat gradient has been suggested by meteorite data all along (40) but, because such planetesimals should also be rich in C, H, noble gases, and other volatiles, one would expect a strong correlation of these volatiles with FeO, which is not observed. Perhaps the FeO content was established in protoplanets (41) rather than in the solar nebula, but then one would expect little correlation with heliocentric distance. A final, speculative possibility is that FeO was inherited mainly from a presolar stage.

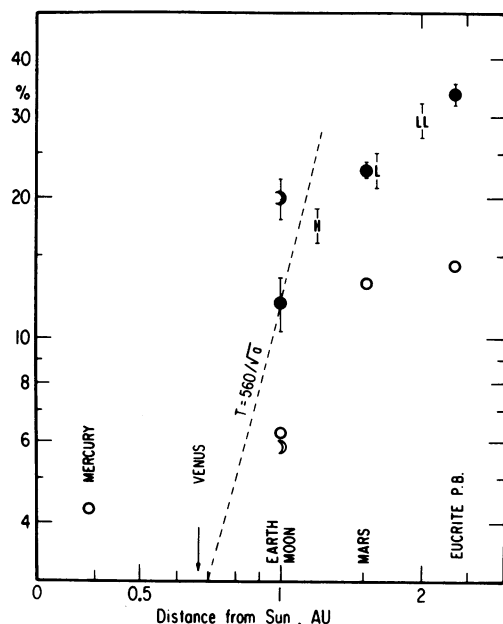


FIG. 5. Both the surface and bulk FeO contents of planets (35) decrease with heliocentric distance a but less steeply than expected from the Fe/FeO equilibrium (22) and an $a^{-0.5}$ dependence of T (dashed line). The letters H, L, and LL designate the corresponding classes of ordinary chondrites. O, Fe^{2+3+} , surface; ●, $\text{FeO}/(\text{FeO} + \text{MgO})$, bulk.

MODEL COMPOSITIONS OF THE EARTH, VENUS, AND MERCURY

The compositions, calculated as in Morgan *et al.* (6), are given in Tables 2 and 3. For Earth, various lines of evidence supporting the previous model should equally apply to the present one: an empirical estimate of bulk composition (18), the Rb/Sr ratio inferred from the isotopic composition of Sr in surface rocks (20, 42), mantle abundances of F, Cl, Br, Rb, and Ba (19) and Ni and Ge (43). For Venus and Mercury, no tests are possible yet in the absence of geochemical data.

Mineral compositions for >30 kb correspond to garnet lherzolite (Table 4). Less than half of the thin Mercurian mantle

Table 3. Model compositions of Earth, Venus, and Mercury

	Earth	Venus	Mercury
Mantle + crust, %	67.6	68.0	32.0
SiO ₂ , %	47.9	49.8	47.1
TiO ₂ , %	0.20	0.21	0.33
Al ₂ O ₃ , %	3.9	4.1	6.4
Cr ₂ O ₃ , %	0.9	0.87	3.3
MgO, %	34.1	35.5	33.7
FeO, %	8.9	5.4	3.7
MnO, %	0.14	0.09	0.06
CaO, %	3.2	3.3	5.2
Na ₂ O, %	0.25	0.28	0.08
H ₂ O, %	0.21	0.22	0.016
K, ppm	200	221	69
U, ppm	0.021	0.022	0.034
Th, ppm	0.076	0.079	0.122
Core, %	32.4	32.0	68.0
Fe, %	88.8	88.6	93.5
Ni, %	5.8	5.5	5.4
S, %	4.5	5.1	0.35
Co, %	0.27	0.26	0.25
P, %	0.62	0.58	0.57

Table 4. Normative mineralogical composition of Earth, Venus, and Mercury

	Earth 30 kb	Venus 30 kb	Mercury	
			30 kb	0 kb
Mantle, wt %	67.6	68.0	32.0	32.0
Chromite	0.94	0.88	3.34	3.34
Ilmenite	0.24	0.27	0.42	0.42
Jadeite	1.78	1.95	0.61	—
Plagioclase				16.52
Alkali feldspar				0.76
Anorthite				15.76
Ca-rich pyroxene	11.76	12.14	18.97	6.37
Hedenbergite	1.45	0.86	0.60	0.20
Diopside	10.31	11.27	18.37	6.17
Ca-poor pyroxene	33.22	39.90	24.02	36.01
Ferrosilite	4.10	2.84	0.76	1.14
Enstatite	29.12	37.06	23.26	34.87
Olivine	37.82	30.30	27.43	37.35
Fayalite	4.67	2.16	0.87	1.19
Forsterite	33.15	28.14	26.56	36.16
Garnet	14.22	14.56	25.21	—
Almandine	1.75	1.04	0.80	—
Pyrope	12.47	13.53	24.41	—
Core, wt %	32.4	32.0	68.0	
Metal	75.33	86.10	99.05	
Fe	68.86	79.74	92.85	
Ni	5.61	5.52	5.38	
Co	0.26	0.26	0.25	
P	0.59	0.58	0.57	
Troilite	24.68	13.90	0.96	
Fe	15.68	8.83	0.61	
S	9.00*	5.07	0.35	

Mantle values in mol % and core values in weight %; each normalized to 100%.

* Density data for the core may also be satisfied by some mixture of S and O rather than by S alone. One possible combination is S = 4.51% (corresponding to a cosmic S/K ratio) and O = 4.49%, leaving $\text{Fe}_{\text{metal}} = 61.01\%$, $\text{Fe}_{\text{troilite}} = 7.58\%$, and $\text{Fe}_{\text{oxide}} = 15.68\%$.

attains this pressure, and at 0 kb the assemblage is plagioclase lherzolite. For the Earth, the 30-kb composition resembles the more fertile natural garnet lherzolites but has less olivine. Most garnet lherzolites are probably partly residual and may be richer in olivine than is primitive mantle material (44). The 30-kb assemblage for Venus resembles that of the Earth except for orthopyroxene < olivine. This would be unusual for an equivalent terrestrial rock and reflects a high Si/(Fe + Mg), resulting from low FeO in the Cytherean mantle. On Mercury, the high refractory content leads to abundant garnet (30 kb) and plagioclase (0 kb); in corresponding terrestrial rocks these generally occur only as accessory minerals.

Planetary mantle mineral compositions derived from averaged compositional estimates (35) in the system CaO-MgO-Al₂O₃-SiO₂ suggest two assemblage sequences (45). Earth, Venus, and the Moon yield plagioclase lherzolite (0–10 kb), spinel lherzolite (10–25 kb), and garnet lherzolite (>25 kb), whereas Mars and Mercury give spinel plagioclase wehrlite, spinel lherzolite, and spinel-garnet wehrlite at corresponding pressures. Using our compositional data, however, we find that all five planets follow the lherzolite sequence, except for Mars, which enters the garnet-wehrlite field at >25 kb.

Now that model compositions are available for six planets, let us look for trends (Fig. 6). Compositions are represented by the five indices from Table 1, poorly known values being indicated by open symbols. The observed range for chondrites is also shown, to check our premise that chondrites and planets were made by the same processes. Two first-order trends are

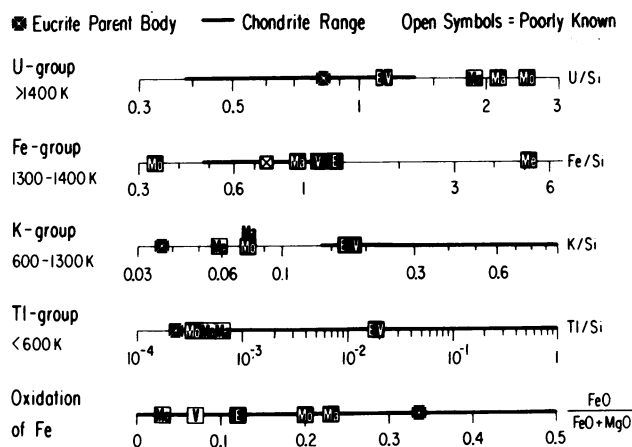


FIG. 6. Chemical composition of planets. Planet composition shows little correlation with heliocentric distance and only a gross correlation with mass: relative to Earth and Venus, the four small planets are depleted in volatiles but (except for the eucrite parent body) enriched in refractories. The small planets also show more extreme fractionation, often falling outside the chondritic range. E, Earth; V, Venus; Ma, Mars; Me, Mercury; Mo, Moon.

the decline of FeO/FeO+MgO with heliocentric distance and, for the four other indices, the dichotomy between the two large planets—Earth and Venus—and the three small ones—Mars, Mercury, and the Moon. The latter show more extreme fractionations: enrichments (U group), depletions (K group, Ti group), or both (Fe group). Beyond that, there are no clear trends, although the generally low position of the smallest and most remote body—the eucrite parent body—suggests that some further regularities may be hidden in the data. However, it seems unlikely that the data will ever fall into a neat, monotonic sequence according to mass or heliocentric distance. As pointed out before (7, 8), some chondrite classes and shergottites—an achondrite class resembling eucrites—match or surpass the Earth in volatile content (46, 47), although they are derived from smaller and more remote bodies. Apparently, mass and heliocentric distance were not the only factors controlling planet composition.

The dichotomy between large and small planets also shows up in the comparison with chondrites. Values for Earth and Venus always fall within the chondritic range, whereas values for the small planets often fall outside. This does not necessarily imply a whole new set of processes because the “chondritic range” is based on a manifestly incomplete sampling of the asteroid belt. And the very factors that caused planets to grow beyond asteroidal size may also have driven the fractionation processes to greater extremes [e.g., enrichment of U group by preferential accretion of early condensate (24)].

The strongest case for a wholly different process can be made for the K group where the planets lie at or below the low end of the chondritic range and for the Fe group where the most extreme values are shown by the Moon (the only satellite) and by Mercury (the planet closest to the Sun). Some authors have explained these bodies by special processes: the Moon, by fission (48) or by collisional differentiation of planetesimals, enriching brittle silicates over ductile nickel-iron (49); and Mercury, by arrested condensation at ≈ 1400 K (Fig. 1) or by gas drag (50). We prefer to invoke the same mechanism (ferromagnetism?) that was responsible for Fe/Si fractionations among chondrites and the remaining planets. The choice obviously depends on one's faith in Occam's razor.

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